6

# IONIZATION DISTRIBUTION IN THE F-REGION

By B. CHATTERJEE*

INSTITUTE OF RADIO PHYSICS AND ELECTRONICS, UNIVERSITY OF CALCUTTA

*(Received for publication, January 12, 1951)*

**ABSTRACT**. The height distribution of ionization in the composite F-region is calculated on the assumptions that the scale height (*i.e.*, the temperature) increases and the recombination coefficient decreases with height in this region. The recombination coefficient is assumed to vary in terms of the reduced height, that is the height (above $F_{1\,max}$) measured in terms of the scale height and not in terms of the actual height as had been done earlier by A. P. Mitra in making similar calculations. The present assumption automatically takes into account the effect of rising temperature on the value of the recombination coefficient Height distribution curves for high and low latitude stations are drawn for typical summer and winter conditions in the high atmosphere in the F-region. It is found that when the possible effects of ionospheric tidal drifts are taken into account, the forms of the distribution curves, as also the seperation between the $F_1$ and $F_2$-layers, agree with those deduced from observational data. The investigation lends support to the hypothesis that the ionospheric regions $F_1$ and $F_2$ belong to a common bank of ionization produced by a common ionizing radiation from the sun.

## 1. INTRODUCTION

According to current ideas, the higher regions of the ionosphere, $F_1$ and $F_2$, owe their ionizations to a common ionizing radiation from the sun (the radiation ionizing O atom at its first ionization potential). It is supposed that owing to the peculiar physical characteristics of the atmosphere prevailing in these high regions and/or due to air motions resulting from tidal forces, two ionization maxima, instead of the usual single one due to the Chapman process, are produced in a single bank of ionization. The physical characteristics that lead to this double maxima are two-fold, namely, increase of temperature (*i.e.*, scale height) and rapid decrease of recombination coefficient with height from above the $F_1$-maximum. These may be compared to the two opposite effects, *viz.* decreasing intensity of ionizing radiation and increasing density of ionizable particles with depth of penetration of the solar rays which lead to the production of the Chapman maximum of ionization. This idea regarding the process of bifurcation of the F-region into $F_1$ and $F_2$ was first suggested by Mohler (1940) and was later extended by Bates and Massey (1946).

The tidal hypothesis for the production of the double maxima was put forward by Martyn (1947). According to this hypothesis the $F_2$-maximum is produced above the normal $F_1$-maximum by the vertical motion of ions under the influence of geomagnetic field. However, though the effects of

* Communicated by Prof. S. K. Mitra.

tides is considered to be significant in modifying the parameters of the $F_2$-region, it does not appear to be large enough to produce the two maxima.

As a test of the first hypothesis one may examine quantitatively if the observed rates of decrease of recombination coefficient and increase of temperature with height do actually lead to the bifurcation process. A closer test would be to see if the various phenomena associated with bifurcation e.g., diurnal, seasonal and latitudinal variations of the $F_1$-$F_2$ separation follow as necessary consequences of the probable height variation of these two parameters. The simpler first-mentioned test was carried out by A. P. Mitra (1952). He made a detailed study of the doubling process of the F-region by taking the effects of the changing scale height and changing recombination coefficient into account. His formula, however, gave a steady increase of ionization density with height in the $F_2$-region under equilibrium condition (i.e., for $dN/dt=0$). He, therefore, argued that the equilibrium condition is seldom attained in this region and introduced a correction factor for the same. When this was done, a second ionization maximum corresponding to the $F_2$-layer was obtained.

In the present paper the hypothesis is examined anew with a somewhat modified assumption regarding the nature of the height variation of the recombination coefficient ($\alpha$). It is assumed that $\alpha$ is a function of the reduced height $z\left(=\frac{h-h_0}{H}\right)$ rather than simply of the height $h$ as assumed by A. P. Mitra. This assumption makes the initial rate of fall of $\alpha$ with height much sharper and automatically takes into account the effects of varying temperature. Further, it is found that when the possible effects of ionospheric tidal drifts are superposed, the forms of the distributions, as also the observed variations of the $F_1$-$F_2$ separation, agree with the observed data.

The assumptions made regarding the variations of temperature and recombination coefficient with height are based on observed data and discussed in the following section.

## 2. ASSUMPTIONS MADE REGARDING THE PHYSICAL PROPERTIES OF THE UPPER ATMOSPHERE AT THE F-REGION HEIGHT

Evidences, partly direct and partly indirect e.g., works of Martyn and Pulley (1936) and rocket data (Mitra, 1952, p. 549) indicate that the upper atmospheric temperature begins to rise from a height of about 85 km, and that the rise continues up to about 400 km, after which the temperature falls gradually with height, to merge itself with the interstellar temperature. The exact nature of the temperature rise with height is not known but it is generally assumed to be linear, e.g., the standard temperature distribution adopted by the National Advisory Committee for Aeronautics, U.S.A. and the temperature distribution in the 'model atmosphere' as assumed by Gerson (1951). Of course, all the methods of temperature determination in the $F_2$-region are indirect and the conclusion of Maeda and Tukada, "It, therefore, seems that the theory involving the temperature of the $F_2$-region is unnatural", can hardly be refuted. But in the present state of our knowledge and technique we have to be satisfied with these results.

We may, therefore, write, for the temperature $T$ at any height $h$ as,

$$T = T_0 + bh$$

where $T_0$ is the temprature at the reference level ($h=0$) and $b$ is the height gradient.

Since the scale height is defined by $H = KT/mg$, one may also write,

$$H = K(T_0 + bh)/mg = H_0 + \beta h. \qquad \ldots \quad (1)$$

where

$$H_0 = KT_0/mg$$

and $\beta$ is the scale height gradient (The variation of $g$ with height is neglected). $T_0$ and $b$ vary with the hour of the day and the season of the year and we shall adopt the distributions as given by Gerson. It may be mentioned in this connection that the value of $T$ (and hence of $H$), as adopted in the model atmosphere of Gerson, was obtained not only from electron concentration but also from collisional frequency and luminosity curves and filter photographs of northern aurorae. Thus one is justified in assuming these values for finding the electron distribution in the F-region.

Regarding the variation of recombination coefficient in the F-region, ionospheric observations show that it is much less (about one order) in the $F_2$-region than in the $F_1$-region. It is also satisfactory to note that according to the theoretical expressions for the recombination coefficient, or rather the effective recombination coefficient ($\alpha$) that have been suggested $\alpha$ will decrease with height. For example, we have two expressions for $\alpha$, both due to Bates and Massey (1946, 1950).

$$\alpha = \alpha_e + \lambda\alpha_i$$

where

$$\lambda = \frac{Bn}{\gamma_1 B + \alpha_i N_e}$$

and

$$\alpha = \frac{\alpha(d) n(O)}{N_e}$$

In the above expressions,

$B$ = coefficient of attachment of electrons to neutral atoms and molecules ;

$n$ = number density of neutral particles ;

$N_e$ = number density of electrons ;

$BN$-$\gamma_1$ = rate of loss of negative ions due to photo-detachment of electrons ;

$\alpha_i$ = coefficient of mutual neutralisation of positive and negative ions ;

$\alpha(d)$ = coefficient of dissociative recombination ;

$n(O)$ = number density of oxygen atoms.

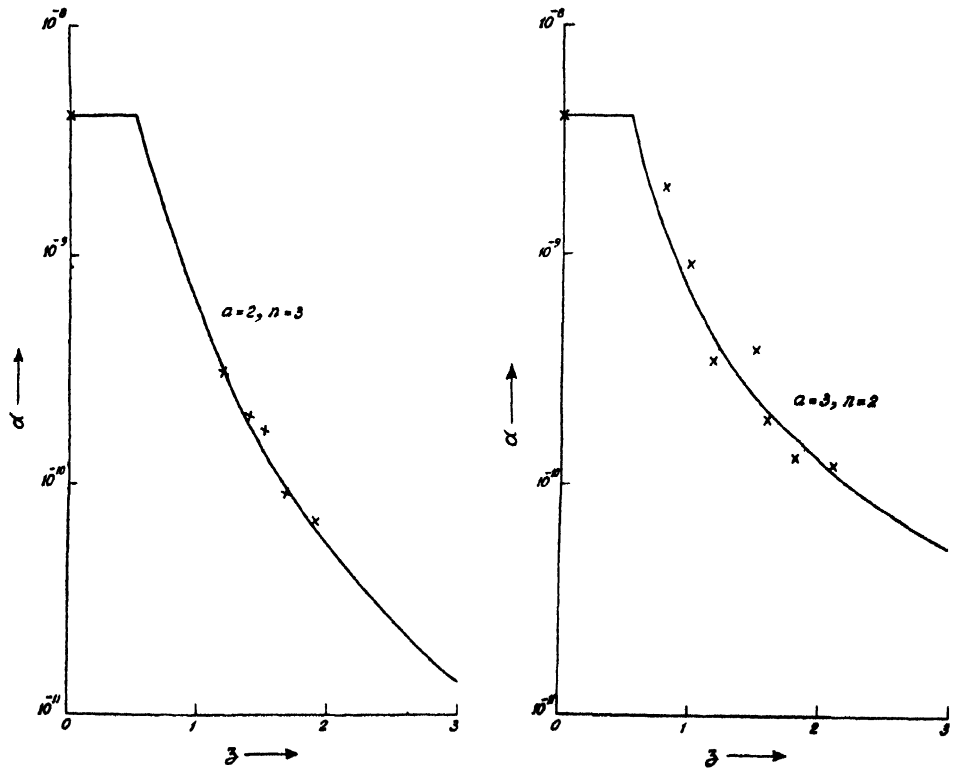
It is seen that according to the first expression, $\alpha$ diminishes with height because $n$ decreases and $N_e$ increases with height in the $F_2$-region. The same is also true for the second expression because $n(O)$ decreases and $N_e$ increases with height.

It may also be mentioned that according to some investigators, the variation of $\alpha$ in the $F_2$-region may be attributed to the height variations of temperature and pressure in this region (Baral and Mitra, 1951)

On account of these complicating factors, it is not possible, in the present state of our knowledge, to give any law of height variation of $\alpha$ that may be strictly acceptable on theoretical grounds. One has, therefore, to be satisfied with empirical relations as may be considered to represent best the observed facts of height variation. We have adopted in our calculation the relation

$$\alpha = \frac{\alpha_0}{(1+az)^n}$$

where $\alpha_0$ is the value of $\alpha$ in the $F_1$-region where it is known to be sensibly constant with height. The multiplication factor $a$ and the power index $n$ are so chosen as to yield approximately the observed variation of $\alpha$ with height. It may be mentioned that A. P. Mitra also took a similar power law of variation




Figs. 1(a) and (b). Variations of $\alpha$ with $z$ for high and low latitude stations as assumed in the calculations. The crosses indicate the observed values : (a) for low latitude station, Calcutta and (b) for high latitude station, Slough. It will be noticed that the curve for the assumed parameter values, namely, $a=2$, $n=3$, for curve (a) and $a=3$, $n=2$ for curve (b) approximately pass through the mean positions of the observed values.

tion in his calculations, the exponent ($n$) being assumed by him to lie between 2 and 3. In our calculations, we have taken $n=3$ and $a=2$ for low latitude stations and $n=2$ and $a=3$ for high latitude stations. The calculated height variation of $\alpha$ with these values of $a$ and $n$, resemble approximately the mean observed variation of $\alpha$ in the regions concerned. This is shown in figures 1 ($a$) and 1($b$). The observed values of the parameters used in calculating the experimental points in the graphs are shown in Tables I($a$) and I($b$).

Calcutta (low latitude) data

| Month and year | $\alpha$ | $h_m$ | $H$ | $z$ |
|---|---|---|---|---|
| January, 1953 | $2\times10^{-10}$ | 300 | 70 | 1.4 |
| February, 1953 | $1.7\times10^{-10}$ | 330 | 87 | 1.5 |
| August, 1946 | $7\times10^{-11}$ | 388 | 100 | 1.88 |
| September, 1946 | $9.2\times10^{-11}$ | 365 | 97 | 1.7 |
| November, 1952 | $3\times10^{-10}$ | 300 | 84 | 1.2 |

TABLE 1($b$)

Slough (high latitude) data

| Month and year | $\alpha$ | $h_m$ | $H$ | $z$ |
|---|---|---|---|---|
| February, 1951 | $3.5\times10^{-10}$ | 275 | 63 | 1.18 |
| April, 1951 | $3.9\times10^{-10}$ | 290 | 60 | 1. |
| May, 1951 | $2\times10^{-10}$ | 315 | 72 | 1.6 |
| June, 1951 | $1.35\times10^{-10}$ | 360 | 84 | 1.91 |
| July, 1951 | $1.25\times10^{-10}$ | 365 | 80 | 2.10 |
| September, 1951 | $3.9\times10^{-10}$ | 275 | 62.5 | 1.2 |
| October, 1951 | $9\times10^{-10}$ | 255 | 54 | 1.01 |
| December, 1951 | $2\times10^{-9}$ | 235 | 43 | 0.8 |

$\alpha$ for 200 km has been taken as $4\times10^{-9}$ (Mitra, 1952, p 291). Cases for which $h_m$, the height of maximum ionization density is less than 400 km have only been considered. The simplified relation, as used in the calculations, does not hold good for $h_m > 400$ km. due to the fall of temperature above this height

The calculations for the Slough data and some of the Calcutta data have been done by the author using the well known method of Appleton. Other data of Calcutta have been taken from the calculations of Baral and Mitra (1951) In Appleton's methods of calculation we have,

$$q=\frac{N_B{}^2\dfrac{dN_A}{dt}-N_A{}^2\dfrac{dN_B}{dt}}{N_B{}^2-N_A{}^2}$$

for values of $q$ for hours, equally spaced on the two sides of local noon. The subscripts $A$ and $B$ give the values for these two equally spaced hours.

Knowing the zenith angle of the sun, the value of $q$ for noon is calculated and then $\alpha$ for noon is obtained from the formula,

$$\alpha = \frac{q - \dfrac{dN}{dt}}{N^2}$$

Now if we assume that $z_0$ is the height up to which the $F_1$-region condition prevails, (*i.e.* $\alpha$ remains sensibly constant) we get,

$$\left.\begin{aligned} &\alpha = \alpha_0 \ ; \ [\text{for } z < z_0] \\ \text{and} \qquad &\alpha = \frac{\alpha_0}{[1 + a(z - z_0)]^n} \ ; \ [\text{for } z > z_0] \end{aligned}\right\} \qquad \ldots \quad (2)$$

In the analysis to follow, only the noon-time condition (for $dN/dt = 0$) has been considered. This is because analytical solution is possible for this case only. Analysis for the general case has not been attempted because the calculations would have been too laborious, involving successive approximation. This study has been reserved for a future communication.

## 3. STRUCTURE OF THE COMPOSITE F-REGION

In the ideal case of an isothermal atmosphere with constant recombination coefficient and monochromatic ionizing radiation, the distribution of ionization with height under equilibrium condition is given by the well known Chapman (1931) formula,

$$N = N_0 e^{\frac{1}{2}(1 - z - \sec \chi\, e^{-z})} \qquad \ldots \quad (3)$$

where $N_0$ = maximum ionization density (for $z = \chi = 0$) formed at a height $h_0$.

$N$ = ionization density at any height $h$, measured from the reference level.

$\chi$ = solar zenith angle.

This formula, however, is not applicable to regions above the E-region because both scale height (i.e. temperature) and recombination coefficient vary with height.

In the major portion of the $F_1$-region, though $H$ varies appreciably with height, the recombination coefficient does not do so. In this case if $\alpha$ is regarded as constant, we can write the variation of $N$ with height, after Nicolet (1951) as

$$N = N_0 e^{\frac{1+\beta}{2}(1 - z - \sec \chi\, e^{-z})} \qquad \ldots \quad (4)$$

Now, the time rate of change of ionization, when the tidal effect is neglected, is given by

$$\frac{dN}{dt} = q - \alpha N^2 \qquad \ldots \quad (5)$$

where, $q$ = the rate of ion production at a height $h$, the variation of which with height is (according to Nicolet)

$$q = q_0 \epsilon^{(1+\beta)(1-z-\sec\chi\, e^{-z})} \tag{6}$$

For equilibrium condition (near noon) $\frac{dN}{dt} = 0$. Hence, $N = \sqrt{q/\alpha}$.

Substituting the values of $q$ from (6) and $\alpha$ from (2) (for $z > z_0$), in the above expression,

$$N = \sqrt{\frac{q_0}{\alpha_0}[1 + a(z - z_0)]^n} \cdot \epsilon^{\frac{1+\beta}{2}(1-z-\sec\chi\, e^{-z})}$$

$$= N_0[1 + a(z - z_0)]^{n/2} \cdot \epsilon^{\frac{1+\beta}{2}(1-z-\sec\chi\, e^{-z})}$$

where, $N_0 = \sqrt{q_0/\alpha_0}$

Or

$$\frac{N}{N_0} = [1 + a(z - z_0)]^{n/2} \cdot \epsilon^{\frac{1+\beta}{2}(1-z-\sec\chi\, e^{-z})} \quad \ldots \quad (7)$$

Hence, for the ionization distribution of the composite F-region,

$$\frac{N}{N_0} = \epsilon^{\frac{1+\beta}{2}(1-z-\sec\chi\, e^{-z})} \; ; [\text{for } z < z_0] \quad \ldots \quad (8a)$$

and

$$\frac{N}{N_0} = [1 + a(z - z_0)]^{n/2} \cdot \epsilon^{\frac{1+\beta}{2}(1-z-\sec\chi\, e^{-z})} \; ; [\text{for } z > z_0] \quad \ldots \quad (8b)$$

The actual value of $z_0$ is a matter of guess-work. It must be positive *i.e.* above normal $F_1$-maximum level ($h_0$) but cannot be very large ; for, we know from experimental observations that the $F_2$-region conditions begin from only a little distance above $h_0$. Considering these, $z_0$ has been taken as 0.5, which is quite reasonable.

In plotting the ionization distribution curve with height ($h$), it is to be noted that,

$$z = \frac{h - h_0}{H} = \frac{h - h_0}{H_0 + \beta h}$$

For simplifying the calculations, let the reference level be at $h_0$, from which the height is to be measured *i.e.* let us put $h_0 = 0$.

Then,

$$z = \frac{h}{H_0 + \beta h} \; ; \text{ or } h = \frac{H_0 z}{1 - \beta z}$$

where $H_0$ is the scale height at $h_0$.

As $h_0$ is at $F_{1\,max}$, $z$ represents height in terms of scale height measured from the maximum of $F_1$. Now, the $F_1$-maximum is formed at height of about 200 km and from experimental observations it is known that the value of $\alpha$ (Mitra, 1952, page 553) at 200 km (*i.e.* $H_0$) is about 50 in summer and 35 in winter. With these data the height variation of ionization can be represented directly in terms of $h$, as has been done in figures 2-5.

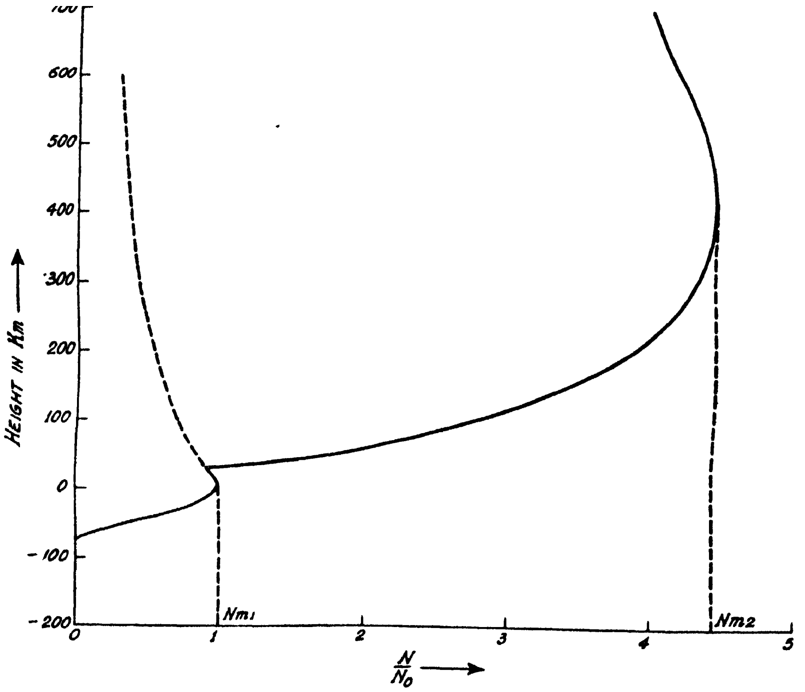



Fig. 2 Variation of ionization density with height as obtained by putting $H_0=50$ km, $\beta=0.25$, $a=2$, $n=3$ in Eqn. (8a) and 8(b). The curve represents a typical ionization distribution at high latitudes in summer noon.

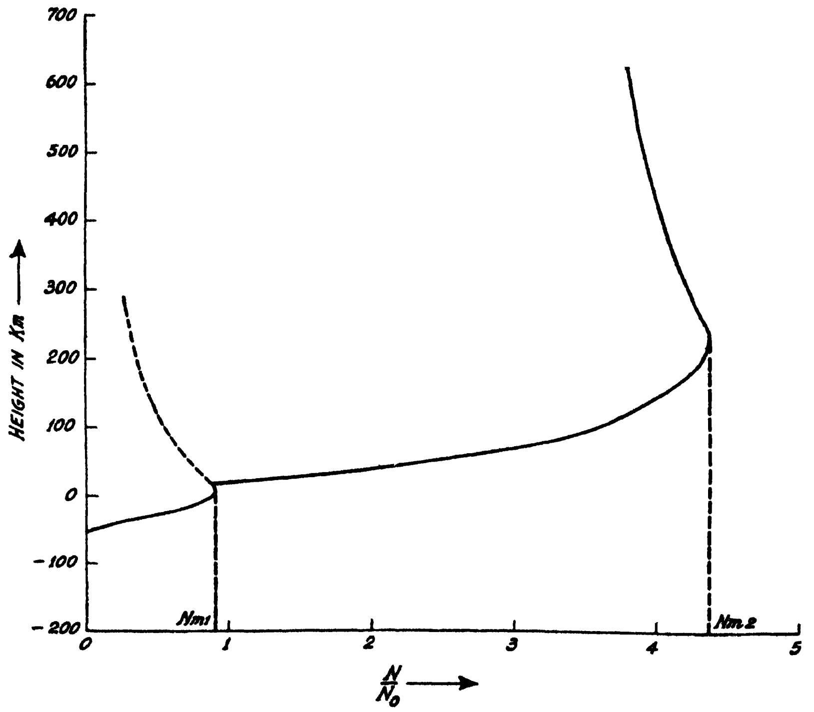



Fig. 3. Variation of ionization density with height as obtained by putting $H_0=35$ km, $\beta=0.2$, $a=2$, $n=3$ in Eqs. (8a) and 8(b). The curve represents a typical ionization distribution at high latitudes in winter noon.

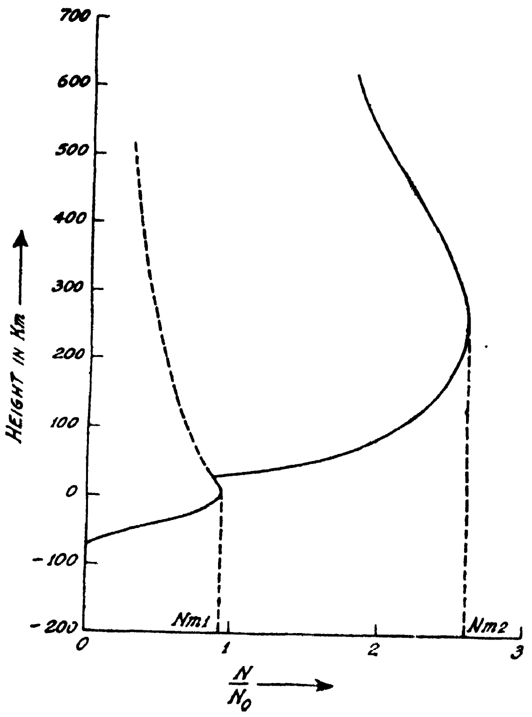



Fig. 4 Variation of ionization density with height as obtained by putting $H_0=50$ km, $\beta=0.25$ $a=3$, $n=2$ in Eqs. (8a) and (8b). The curve represents a typical ionization distribution at low latitudes in summer noon.

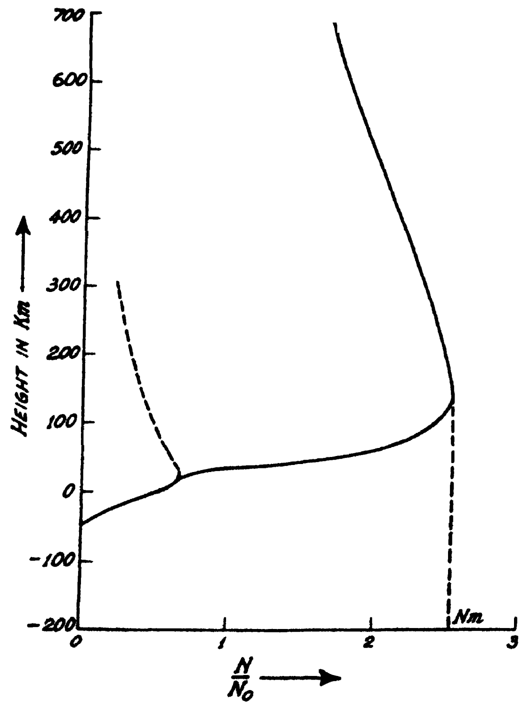



Fig. 5. Variation of ionization density with height as obtained by putting $H_0=35$ km, $\beta=0.2$, $a=3$, $n=2$ in Eqs. (8a) and (8b). The curve represents a typical ionization distribution at low latitudes in winter noon.

## 4. IONIZATION DISTRIBUTION WITH HEIGHT

1. *For low latitudes* :

(*a*) *Summer noon condition*, $\chi=0°$ (figure 2). We assume the value of $\beta$ to be 0.25. For the adjustable parameters $a$ and $n$, we assume values $a=2$, $n=3$. As already mentioned these values of $a$ and $n$ are chosen to suit approximately the observed height variation of $\alpha$ at low latitude stations. The figure shows two distinct maxima of ionization. The $F_2$-layer has a very large semi-thickness and the $F_{2\ max}$ is situated at a height of about 400 km from the $F_{1\ max}$. The results correspond well with experimental observation, particularly when the tidal effects are taken into account.

(*b*) *Winter noon condition*, $\chi=30°$ (figure 3). In this case the values of $a$ and $n$ are kept the same as in figure 2. Only a lower value of $\beta$ $(=0.20)$ is taken, as the temperature gradient is lower in winter. In this case also two distinct maxima in ionization distribution are observed. But the semi-thickness of the $F_2$-region has become much smaller—$F_{2max}$ being situated at a height of about 200 km from the $F_{1\ max}$. This lowered height of $F_{2\ max}$ is due to decreased temperature in winter. It is also to be noted that the winter height of the $F_1$-maximum is situated about 10 km above the summer height. Thus we see that though the height of $F_{2\ max}$ changes by a large amount from summer to winter that of $F_{1\ max}$ remains practically the same. The ionization density ratio $\frac{N_{m2}}{N_{m1}}$ is, however, much larger in winter than is shown in the figure. This large value of $N_{m2}$ in winter is due to tidal effects as explained qualitatively in the next section.

2. *For middle and high latitudes* :

(*a*) *Summer noon condition*, $\chi=20°$ (figure 4). The value of $\beta$ is assumed to be equal to 0.25 as in figure 2. The value of $a$ and $n$ are, however, taken respectively to be 3 and 2. This makes the height variation of recombination co-efficient slightly less sharp than its value in low latitudes, as is supposed to be the case. Here also the $F_2$-region is very thick, and the $F_{2\ max}$ is formed at a height of about 250 km above $F_{1\ max}$. $\frac{N_{m2}}{N_{m1}}$ is 2.8. All these results agree well with the experimental observations in middle and slightly higher latitudes, specially when the tidal corrections are made.

(*b*) *Winter noon condition* ; $\chi=60°$ (figure 5). As in the case of low latitude stations, the value of $\beta$ is taken to be 0.2, a little lower than that of the summer noon value. The values of $a$ and $n$ are taken to be the same. It is to be noted that the normal $F_1$-maximum (under the condition of constant recombination coefficient) lies above the point $z_0$ due to the large value of $\chi$. There is only one maximum in the ionization distribution curve of the composite F-region formed at a height of about 140 km above $h_0$. This is what is observed in high latitude stations in winter. From this it may be concluded that if the solar zenith angle is larger than a certain given value, so that the normal $F_1$-maximum occurs above the point $z_0$, the actual value of $z_0$ depending on local conditions—then there will be a single maximum in the F-region at noon *i.e.* there is no bifurcation.

The ionization distribution in the very high latitude stations and in the auroral zones is largely governed by ionizations caused by impact of charged particles and hence it cannot be represented by equations. (8*a*) and (8*b*). Also, on the magnetic equator (*e.g.* at Huancayo), the ionization distribution is rendered complicated by the strong current system (electro-jet) flowing over it. Excepting these special regions, equations (8*a*) and (8*b*) give for other parts of the globe, the noon time ionization distribution more or less correctly, subject to the tidal influences which are discussed below.

## 5. MODIFICATION IN THE IONIZATION DISTRIBUTION DUE TO TIDAL EFFECTS

According to Martyn (1947) equation (5) is modified as follows in presence of tidal effects:

$$\frac{dN}{dt} = q - \alpha N^2 + \frac{d}{dz}(Nv) \qquad \ldots \quad (9)$$

where $v$ is the vertical ionic drift velocity, *measured positively downwards.*

For conditions in the $F_2$-region, Martyn solved this equation with the approximation that $(q - \alpha N^2)$ is negligible compared to $\frac{d}{dz}(Nv)$. A. P. Mitra (1951) has made an accurate analysis for the night time condition *i.e.* for $q = 0$; and recently Weiss (1953) has made a generalised analysis of Eq. (9) under the condition of a constant recombination co-efficient. All these calculations show that tidal effects modify the ionization distribution in the $F_2$-region, as deduced from Eqs. (8*a*) and (8*b*); and that the modifications are such as to make them more in accord with the observed results.

It is to be noted that, on account of the high recombination co-efficient at lower heights, the ionization density of the undisturbed $F_2$-region will decrease when the drift velocity is downwards and will increase when it is upwards. The separation between the two layers ($F_1$ and $F_2$) will thus always increase for downward drift and generally decrease for upward drift. We have used 'generally' for the latter, because if the gradient of drift is steep in the upper part of the $F_2$-region, then an upward drift may cause an increase of separation.

Now, calculations by Martyn show that for low latitude stations in summer, the drift is downwards at noon. As a result of this, the maximum ionization density of the $F_2$-region is decreased, and the separation between $F_1$ and $F_2$, which is already large, increases further. At such latitudes in winter, there is a reversal of phase of the drift velocity somewhere between the $F_1$ and $F_2$-layers. At the level of reversal the amplitude of $v$ must be very small. As such, there is very little modification in the maximum ionization density at this level. The separation between the two regions, however, increases considerably in the winter due to the large downward drift of ions in the upper part of the $F_2$-region.

For high latitudes in summer, the drift velocity is found to be downwards in the forenoon , though, the magnitude of the drift is not very large. This causes some decrease in the maximum ionization density of $F_2$ and increases further the already large separation between $F_1$ and $F_2$. Further, in such latitudes, in winter, there is a small upward drift at noon. This increases the maximum ionization density of $F_2$ and helps the merging together of the two regions, decreasing their separation, if any.

We thus see that as a result of contribution by the tidal drifts, the separation between $F_1$ and $F_2$ at low latitudes is large both in summer and in winter (the summer separation being relatively large due to increased temperature) ; but at middle and high latitudes it is large only in summer (Ghosh, 1953). In winter, the separation is small at middle latitudes. This further diminishes with increase of latitude and ultimately vanishes in high latitudes where the two regions merge together.

## 6. DISCUSSION

Simple inspection of the curves in figures 2-5 shows that when the recombination co-efficient is decreasing and the scale height is increasing with height, then, *if the solar zenith angle is not very large*, a secondary maximum is produced above the normal $F_1$-maximum. But, if the solar zenith angle is very large (as in winter at high latitude stations) only a single maximum is produced in the F-region. It is formed slightly above the height where the normal $F_1$-maximum would have been, and has a larger ionization density.

The above results at once explain the diurnal variation of $F_1$-$F_2$-separation when and where such separation exists. In the morning, as the solar zenith angle is very large, the $F_{1\ max}$ is formed above $z_0$ and hence there is no bifurcation, only one maximum being formed. As the sun moves up in the sky, $F_{1max}$ comes down (below $z_0$) and two distinct maxima are formed. As the solar zenith angle decreases further with the approach of noon, the height of $F_{1\ max}$ goes down further and due to the increase in the temperature above, $F_{2\ max}$ moves up (at a much faster rate). Thus, with advance of the day, $F_1$-$F_2$ separation increases, being maximum at about noon. The reverse process takes place in the afternoon hours when the solar zenith angle increases. Near sunset, $F_{1\ max}$ moves up above $z_0$ and the two layers merge together.

The slight irregularities observed in the diurnal variation of $F_{1\ max}$ are also explained by Eq. (8*a*). From experimental observations it is known that in low and middle latitude stations, the diurnal variation of $F_1$-region critical frequency is approximately given by $(\cos\chi)^{0.8}$, instead of by

$(\cos \chi)^{0.25}$, as given by the Chapman law. This is explained by Eq. (8*a*), according to which the diurnal variation should be given by $(\cos \chi)^{\frac{1+\beta}{4}}$, which, for $\beta = 0.2$ (as taken), is $(\cos \chi)^{0.3}$.

In discussing the results of calculations it should be noted that the curves drawn in figures. 2—5 are only illustrative and not representative of actual conditions prevailing in any particular station. It is also to be mentioned that the effect of fall of temperature with height above 400 km has not been considered in our simplified calculations. The calculated heights of $F_{2\ max}$, which are above 400 km, are to be reduced to certain extent due to the effect of this negative temperature gradient. At very low latitudes (near the equator) both $H_0$ and $\beta$ are high throughout the year. As such, the ionization distribution (neglecting the tidal corrections) should be almost the same throughout the year, analogous to that in figure 2. The seasonal variation of ionization densities and $F_1$-$F_2$ separations in these stations, as observed, are mainly due to tidal effects as explained in the paragraph 4 of the previous section. As pointed out in that paragraph, the tidal effect causes the maximum ionization density of the $F_2$-region to decrease in summer and the $F_1$-$F_2$ separation to increase in both the summer and winter solstices. Also, at high latitudes in winter, $H_0$ is much smaller and $h_0$ is at a lower height. Hence, the single ionization maximum in the F-region is produced at a much lower height, being just above the height where the normal summer-time $F_{1\ max}$ would have been. This lowering is further helped by a small upward tidal drift as discussed in the paragraph 5 of the previous section.

During periods of high solar activity the enhanced solar ionizing radiation increases the rate of ion production ($q$) and raises the temperature level (both $H_0$ and $\beta$) in the F-region throughout the globe (excepting at high latitude stations in winter). As a result, there is increased ionization and increased separation between the $F_1$ and $F_2$-layers during such periods (Ghosh, 1953). (In winter at high latitudes, due to very large zenith angle of the sun and very small daylight hours, the upper atmospheric temperature, in so far as it is controlled by the absorption of solar radiations, is very little affected by solar activity).

It thus appears that Eqs. (8*a*) and (8*b*) derived on the hypothesis that $F_1$ and $F_2$ belong to a common bank of ionization can explain all the so-called 'anomalous' behaviours of the $F_2$-region when proper values of the parameters $H_0$, $\beta$, $a$ and $n$ are introduced in Eqs. (8*a*) and (8*b*) and account is taken of the relevant tidal effects. The equations are applicable to the $F_1$ *cum* $F_2$ conditions at all points on the globe excepting over the auroral regions and over the magnetic equator, for reasons already explained.

ACKNOWLEDGMENTS

The author wishes to express his grateful thanks to Prof. S. K. Mitra for his constant guidance, help and encouragement throughout the progress of the work. Thanks are also due to the Scientific Man Power Committee, Government of India, for financial help.

REFERENCES

Baral, S. S. and Mitra, A. P, 1951, *Jour. Atoms. Terr. Phys.*, **1**, 95.
Bates, D. R., 1950, *Phys. Rev.*, **78**, 492.
Bates, D. R. and Massey, H. S. W., 1946, *Proc. Roy. Soc. A.*, **187**, 24.
Chapman, S., 1931, *Proc. Phys. Soc.*, **43**. 26.
Gerson, N. C., 1951, *Rep. Prog. Phys. Soc. Lond.*, **14**, 316
Ghosh, Mrinmayee, 1953, *Ind. Jour. Phys*,, **27**, 421.
Martyn, D. F., 1947, *Proc. Roy. Soc A.*, **189**, 241.
Martyn, D. F. and Pulley, O. O., 1936, *Proc. Roy. Soc. A.*, **154**, 455
Mitra, A. P., 1952, *Ind. Jour. Phys.*, **26**, 79
Mitra, A. P, 1951, *Jour. Atoms Terr. Phys.*, **1**, 286.
Mitra, S K., 1952, The Upper Atmosphere (2nd Edition), p. 313.
Mohler, F. L., 1940, *Bur. Std. Jour. Res.*, **25**, 507.
Nicolet. M., 1951, *Jour. Atoms. Terr. Phys.*, **1**, 141.
Weiss, A. A., 1953, *Jour. Atoms. Terr. Phys.*, **3**, 30.